MAIN-SHOCK CHARACTERISTICS

BROADBAND STUDY OF THE SOURCE CHARACTERISTICS OF THE EARTHQUAKE

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CONTENTS

ABSTRACT

We have determined the source characteristics of the 1989 Loma Prieta earthquake, using teleseismic data. The solution from body waves gives a mechanism with a strike of 128°, a dip of 70°, a rake of 138°, and a seismic moment of 3×10^{26} dyne-cm (M_w=6.9). This solution is similar to those obtained from long-period Rayleigh and Love waves, P_{nL} waves, and first-motion data. The body-wave solution suggests a depth of about 15 km. The effective duration of the source is 6 s, suggesting lengths of 30 and 15 km for bilateral and unilateral faulting, respectively. Considering the extent of the aftershock zones, we estimate a total rupture length of 35 km. The strike-slip and thrust components of coseismic slip are 177 and 159 cm, respectively. The large thrust component raises an important question regarding the recurrence pattern. If the 1989 Loma Prieta earthquake is a characteristic earthquake with a recurrence interval of about 100 yr, the 159-cm displacement implies a long-term uplift rate of about 1 cm/yr, which appears too high for this region. Three hypotheses for reconciling this apparent conflict are that (1) the geometry of plate motion along the Santa Cruz Mountains section of the San Andreas fault changes on a time scale of several thousand years, and so the coseismic displacement has not accumulated enough to produce high topographic relief; (2) the coseismic-slip direction varies from event to event; and (3) the slip plane of the 1989 Loma Prieta earthquake is distinct from the Pacific-North America plate boundary—if so, then this earthquake is a rather rare, noncharacteristic event. The surface slip of about 1 m for the 1906 San Francisco earthquake is one of the key data in long-term forecasting. No surface slip was observed in the 1989 Loma Prieta earthquake, even if the horizontal slip at depth was as large as 1.8 m. This discrepancy points to a risk of relying too heavily on surface observations for long-term seismic-risk analysis.

INTRODUCTION

The 1989 Loma Prieta earthquake occurred within a seismic gap that had been identified as having a higher than 30 percent (in 30 yr) probability of producing an earthquake of M=6.5–7 (Lindh, 1983; Sykes and Nishenko, 1984; Scholz, 1985; Working Group on California Earthquake Probabilities, 1988). This forecast was based on the historical seismicity and low background seismicity in this gap, and on the amount of surface break (approx 1 m) in the 1906 San Francisco earthquake. Thatcher and Lisowski (1987) argued, however, on the basis of geodetic data, that the coseismic slip for the 1906 earthquake was about 2.6 m, and so it will take more than 150 yr to accumulate this amount of slip (a long-term slip rate of 1.5 cm/yr is assumed for the San Andreas fault in this region), implying that a large earthquake is unlikely in the next few decades.

Now that the 1989 Loma Prieta earthquake has occurred, it is important to assess how it compares with the published forecast. To this end, we have analyzed seismic data, primarily broadband seismograms, to determine the source characteristics of the earthquake. The data are summarized in table 1.

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Table 1.—*Stations and data sets used to determine the source characteristics of the 1989 Loma Prieta earthquake*

Station	Δ (°)	ϕ (°)	ϕ_B (°)	Phases
ARU	86.9	360	0	P, SH, R_1, G_1
OBN	85.4	12	343	P, SH, R_1, G_1
SSB	84.9	35	320	P
HRV	38.6	66	279	P, SH, R_1, G_1
WFM	38.6	66	279	P, SH
ANMO	12.6	95	284	R_1, G_1
CAY	70.7	99	307	P, SH
RPN	64.9	168	349	P, SH, R_1, G_1
PPT	60.3	211	25	P, SH
KIP	35.0	254	56	SH

MECHANISM

Because the most complete data we could obtain are broadband data from International Deployment of Accelerometers/Incorporated Research Institution for Seismology (IDA/IRIS) and GEOSCOPE stations, we first describe the source mechanism calculated from these data. In our analysis, all the seismograms are deconvolved to ground-motion displacements; the data are plotted in figure 1. We used the method of Kikuchi and Kanamori (1989) to invert the records and determine the mechanism. The observed seismograms are matched by synthetic seismograms computed for a sequence of subevents distributed on a fault plane. The Green's functions for five independent moment-tensor elements are computed, and the subevents are represented by a linear combination of these elements.

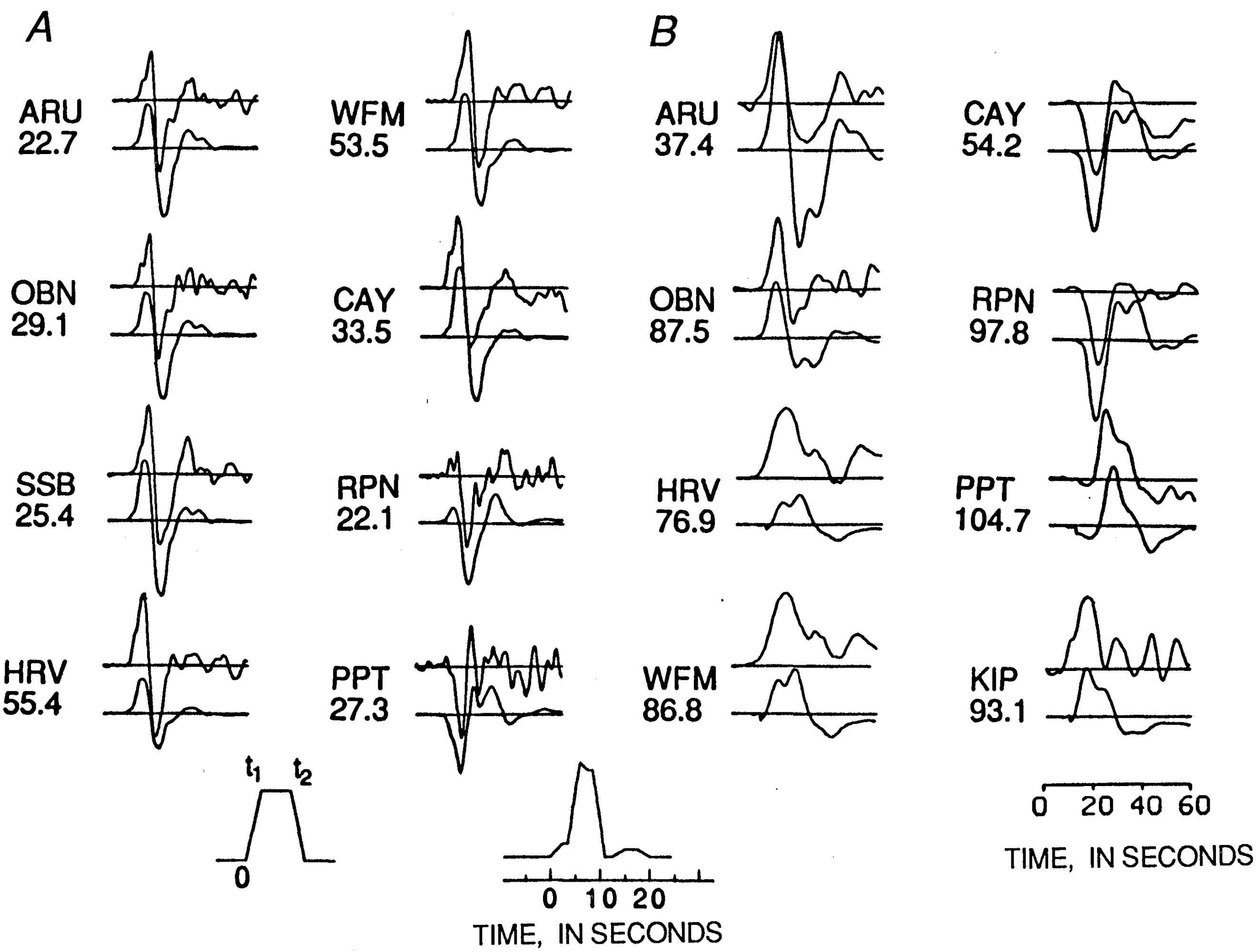

Figure 1.—Observed (top plot) and synthetic (bottom plot) seismograms of *P*-wave (*A*) and *SH*-wave (*B*) ground-motion displacement in 1989 Loma Prieta earthquake. Numbers below three-letter station codes are absolute displacement amplitude (distance from baseline to either peak or trough of observed displacement, whichever is larger) (in microns). Trapezoidal source and time function for three-event source are shown at bottom. Stations are arranged in order of increasing azimuth clockwise from north.

By minimizing the difference between the observed and synthetic seismograms, we determine the moment tensor or mechanism of all subevents, as well as their spatial location and timing.

Because many free parameters are involved in this type of inversion, tradeoffs between different source parameters could occur. First, we use a simplification to obtain the overall model. We assume a single source with a trapezoidal time function (t_1, t_2), as shown in figure 1; we vary t_1 and t_2 to obtain the best solution. We use the parameters listed in table 2 for both the source and receiver structures (common to all the stations). We use an attenuation time constant t^*=1 and 4 s and weights of 3 and 1 for *P* and *S* waves, respectively. We tried three discrete depths, 10, 15, and 20 km, and obtained a best fit at 15 km. The inversion results in t_1=2.5 s, t_2=5 s, a seismic moment (M_0) of 2.4×10^{26} dyne-cm, and a focal mechanism with a strike of 128°, a dip of 70°, and a rake of 138°. Though simple, this source explains the overall features of the observed waveform, and the residual (observed minus synthetic) waveforms are very small.

In the method of Kikuchi and Kanamori (1989), the inversion obtains successive point sources to fit the residual waveforms. In our model, the first point source explains most of the data, and so the other point sources are relatively small.

We assume that the mechanism of all subevents is the same as that of the first subevent, and invert the data. Because later subevents are small and their significance is questionable, we consider the first two or three subevents with a total seismic moment of 2.9×10^{26} or 3.1×10^{26} dyne-cm, respectively. The synthetic waveforms in the three-event model are compared with the observed seismograms in figure 1. Because of noise in the data, especially the significant site response at some stations (for example, KIP), the decision on where to terminate the sequence (iteration in the inversion) is arbitrary. If we include all

Table 2.—*Parameters used for source and receiver structures*

α (km/s)	β (km/s)	ρ (g/cm³)	*H* (km)
5.5	3.18	2.6	4
6.3	3.64	2.67	23.4
6.8	3.93	2.8	5.0
8.0	4.64	3.2	---

subevents, the total seismic moment increases considerably. Considering the total seismic moment calculated from long-period waves (as described below), we judge a seismic moment significantly larger than 3×10^{26} dyne-cm to be unrealistic. In the section below entitled "Coseismic Slip," we use a rounded value of 3×10^{26} dyne-cm for the seismic moment of this subevent; from body-wave data alone, any value between 2.5×10^{26} and 3.5×10^{26} dyne-cm is acceptable. The results are summarized in figure 2 and listed in table 3.

Although the effective duration of the principal subevent is 5 s (fig. 1), we estimate an effective duration of 6 s, allowing for the contributions from smaller subevents.

We use the method of Kawakatsu (1989) and invert long-period surface waves to determine the centroid moment tensor (CMT) of the source. In this inversion, we use both the fundamental-mode and overtone Love and Rayleigh waves. The pass band of the filter is from 3.5 to 7 MHz. The results are shown in figures 2*A* and 2*B* and listed in table 3.

To examine the possible increase in seismic moment at long periods, we invert surface waves at a period of 256 s separately, using the method of Kanamori and Given (1981). In this inversion, the dip angle of one of the nodal planes (70°) is fixed to avoid instability in the inversion;

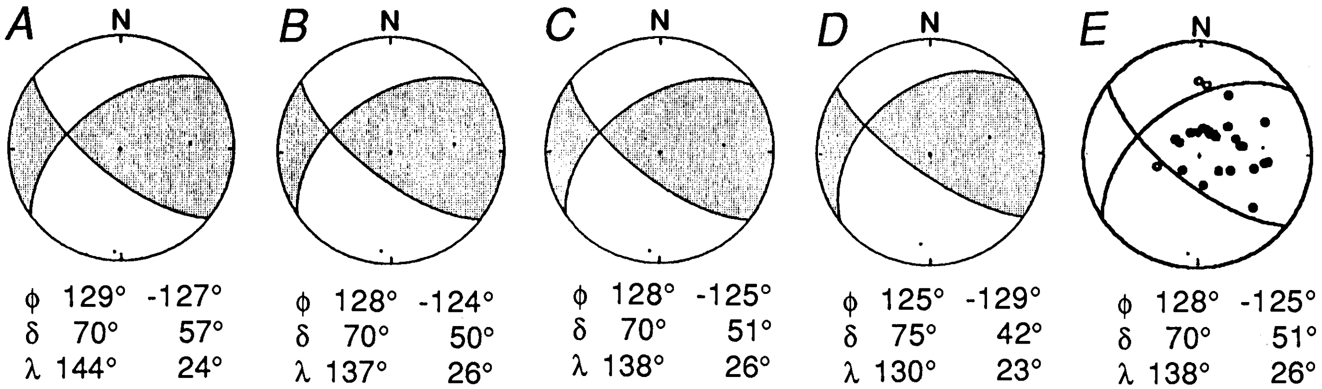

Figure 2.—Equal-area lower-hemisphere projections of focal mechanisms for 1989 Loma Prieta earthquake, arranged in order by decreasing period, obtained from different data sets: (*A*) long-period surface waves, (*B*) centroid moment tensor, (*C*) *P–SH* waves, (*D*) P_{nL} waves, and (*E*) first motions. Nodel-plane parameters: ϕ, strike; δ, dip; λ, rake. Nodal planes for first-motion data (fig. 2*E*) are from solution obtained from *P–SH* waves.

Table 3.—*Source characteristics of the 1989 Loma Prieta earthquake*

Data	M_0 (10^{26} dyne-cm)	ϕ_1 (°)	δ_1 (°)	λ_1 (°)	ϕ_2 (°)	δ_2 (°)	λ_2 (°)
P and *SH* waves-------	3	128	70	138	−125	51	26
Surface waves ---------	2.5	128	70	137	−124	50	26
R_1 and G_1 waves------	2.8	129	[1]70	144	−127	57	24
P_{nL} waves--------------	2.9	125	75	130	−129	42	23

[1]Fixed.

the mechanism is illustrated in figure 2*C*. The estimated seismic moment is 2.8×10^{26} dyne-cm, essentially the same as in the body-wave and CMT solutions. No evidence was found for an increase in seismic moment with period. The first-motion data plotted in figure 2*E* are also consistent with the body-wave and CMT solutions.

Woods and others (in press) used P_{nL} waves recorded at Pasadena, Calif., to determine the source mechanism by matching the waveforms with synthetic seismograms. Their preferred solution is plotted in figure 2*D* and listed in table 3.

The mechanisms obtained from data sets with different periods are essentially the same (fig. 2; table 3), and the seismic moment determined from body waves (period, approx 10 s) is about the same as that determined from surface waves with a period of 256 s, suggesting a simple source for the earthquake. The estimated seismic moment is 3×10^{26} dyne-cm, which gives M_w=6.9.

SOURCE DURATION AND RUPTURE LENGTH

Figure 1 shows that the effective duration of the source is about 6 s, a value that can be used to infer the rupture length. If the rupture is unilateral, then the effective duration τ is given by $\tau=(L/V)-(L\cos\Theta/c)$, where L is the rupture length, V is the rupture velocity, Θ is the azimuth of the station measured from the rupture direction, and c is the body-wave phase velocity along the free surface. Because the *P*-wave phase velocity is much higher than the rupture velocity, the second term is much smaller than the first. Thus, the rupture length L is approximately $L=V\tau$=15 km if τ=6 s and V=2.5 km/s. If the rupture is bilateral, then the rupture length is about twice that for a unilateral rupture—that is, 30 km.

The main shock is located near the center of the aftershock zone (U.S. Geological Survey staff, 1990), suggesting bilateral faulting. If the fault rupture is bilateral, our estimate of the effective duration, (6 s), suggests that L=30 km. The total length of the aftershock zone is about 40 km (U.S. Geological Survey staff, 1990). Although teleseismic data cannot resolve details of the slip distribution on the fault, the rupture length almost certainly does not exceed the length of the aftershock zone. In the following calculations, we use L=35 km, although a shorter rupture length is not precluded.

COMPLEXITY

As shown in figure 1, the source process of the earthquake is simple. The displacement waveform of the earthquake is compared with that of the 1988 Armenia earthquake (M_w=6.7), as recorded at station HRV, in figure 3. Comparison at other stations exhibits essentially a similar difference. The seismograms in figure 3 suggest that the duration of the 1988 Armenia earthquake was 40 s or even longer, in striking contrast to that of the 1989 Loma Prieta earthquake (approx 6 s). Pacheco and others (1989) suggested a substantial variation in focal mechanisms during the first 10 s of the 1988 Armenia earthquake.

This comparison clearly demonstrates the simplicity of the source process of the 1989 Loma Prieta earthquake, which involved a relatively short fault segment. This difference in source complexity may have profoundly influenced the damage potential of this earthquake. Although it is generally agreed that the heavy damage caused by the 1988 Armenia earthquake was primarily due to poor building construction, the source complexity and long duration almost certainly contributed as well. Because source complexity is closely related to local tectonic structure, earthquakes of the same magnitude as the 1989 Loma Prieta earthquake in different tectonic environments can be even more damaging than that earthquake.

COSEISMIC SLIP

If we assume a fault length of 35 km, a fault width 12 km, a rigidity of 3×10^{11} dyne/cm^2, we calculate a coseismic slip of 238 cm from the estimated seismic moment of 3×10^{26} dyne-cm. The strike-slip and thrust components of displacement are 177 and 159 cm, respectively. The average stress drop is estimated at 50 bars.

DISCUSSION AND CONCLUSION

Two aspects of the 1989 Loma Prieta earthquake are noteworthy: the thrust component, as large as 159 cm, and the short rupture length, only 35 km, for an M_w=6.9 earthquake. The large thrust component (1.6 m) raises an important question regarding the recurrence interval of earthquakes along the Santa Cruz Mountains section of the San Andreas fault.

The long-term forecast of this earthquake is based on a combination of the historical seismicity, the estimated slip rate along the Santa Cruz Mountains section of the San Andreas fault, and the surface slip (approx 1 m) in the 1906 San Francisco earthquake (Lindh, 1983; Sykes and Nishenko, 1984; Scholz, 1985). Implicit in this forecast is a relatively short recurrence interval, about 80 to 100 yr. If the 1989 Loma Prieta earthquake is a characteristic event to be expected along this section of the San Andreas fault, with a recurrence interval of about 100 yr, the thrust component of 1.6 m during the earthquake implies a long-term uplift rate of about 1 cm/yr, comparable to the highest rate observed in the world (for example, Yonekura, 1983). An uplift rate this large is generally associated with spectacular topographic relief. Although the long-term uplift rate in the epicentral area is unknown, the regional geomorphology does not seem to reflect such a high rate. We present three hypotheses to reconcile this apparent conflict.

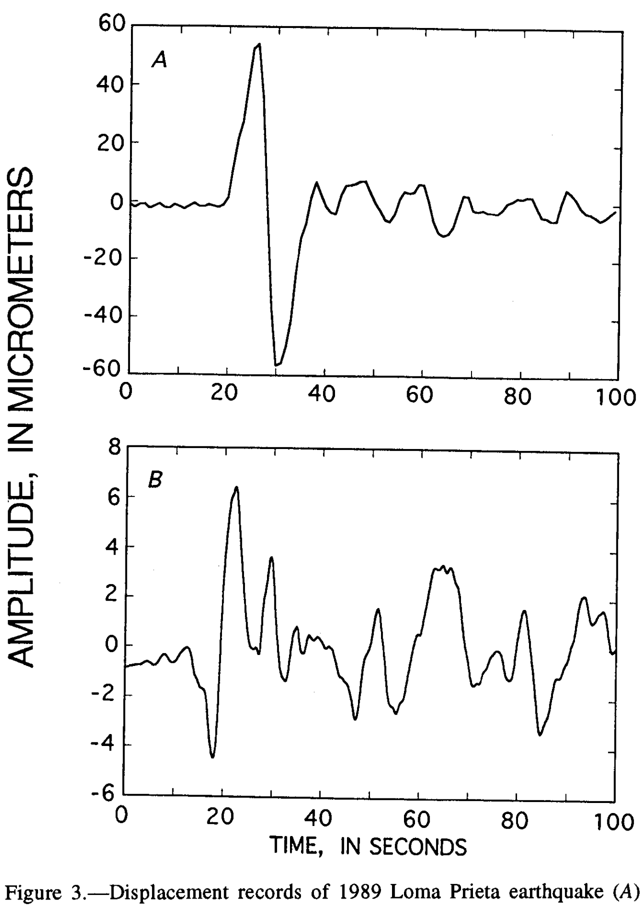

Figure 3.—Displacement records of 1989 Loma Prieta earthquake (*A*) and 1988 Armenia earthquake (*B*) from station HRV.

The first hypothesis is that the geometry of plate motion along the Santa Cruz Mountains section of the San Andreas fault changes on a time scale of several thousand years, and so not enough coseismic vertical displacement has accumulated to produce high topographic relief. If so, then the 1989 Loma Prieta earthquake can be considered a characteristic event along this section of the fault on this time scale.

The second hypothesis is that the coseismic-slip direction varies from event to event. For example, in the 1906 San Francisco earthquake, the motion along this section of the San Andreas fault was essentially strike slip, driven by much larger strike-slip displacements along the adjacent section. Even in the earlier events, which involved the Santa Cruz Mountains section only, the motion could have been primarily strike slip if sufficient stress had not accumulated there to cause vertical displacement.

The third hypothesis is that the slip plane of the 1989 Loma Prieta earthquake is distinct from the Pacific-North America plate boundary. If so, then this earthquake is a rather rare event and not a characteristic event along the San Andreas fault. Although no obvious geologic evidence exists, this conclusion cannot be ruled out.

The short rupture length, about 35 km, of the 1989 Loma Prieta earthquake is highly anomalous, in light of the empirical data available for shallow crustal earthquakes, as shown in figure 4. In fact, the estimated magnitude of the forecasted event is based on the empirical relation plotted in figure 4. For example, Scholz (1985) identified a 75-km-long slip-deficit segment along the Santa Cruz Mountains section of the San Andreas fault and forecast an M=6.9 earthquake. Lindh (1983) identified a 35-km gap and associated it with an M=6.5 earthquake. Both of these estimates are consistent with the empirical relation plotted in figure 4. The anomalous fault-length/moment relation for the 1989 Loma Prieta earthquake is the cause of the discrepancy between the predicted and observed events.

The surface slip of about 1 m during the 1906 San Francisco earthquake is one of the key data in a long-term forecast. Thatcher and Lisowski (1987) argued, however, that the slip of 2.4 m at depth determined from geodetic data should be used for estimating long-term probability. No surface slip was observed in the 1989 Loma Prieta earthquake, even if the horizontal slip at depth was as large as 1.8 m. This discrepancy points to a risk of relying too heavily on surface observations for long-term seismic-risk analysis.

The case for a long-term forecast of the 1989 Loma Prieta earthquake testifies to the importance of synthesizing seismologic, geologic, geodetic, and historical data to obtain probabilistic parameters for long-term hazard assessment and planning. The quantitative analysis of modern seismologic data has revealed many important details, such as source complexity, fault geometry, and rupture length, which, in conjunction with the probabilistic parameters, provide key information for implementing effective seismic-hazard-reduction measures.

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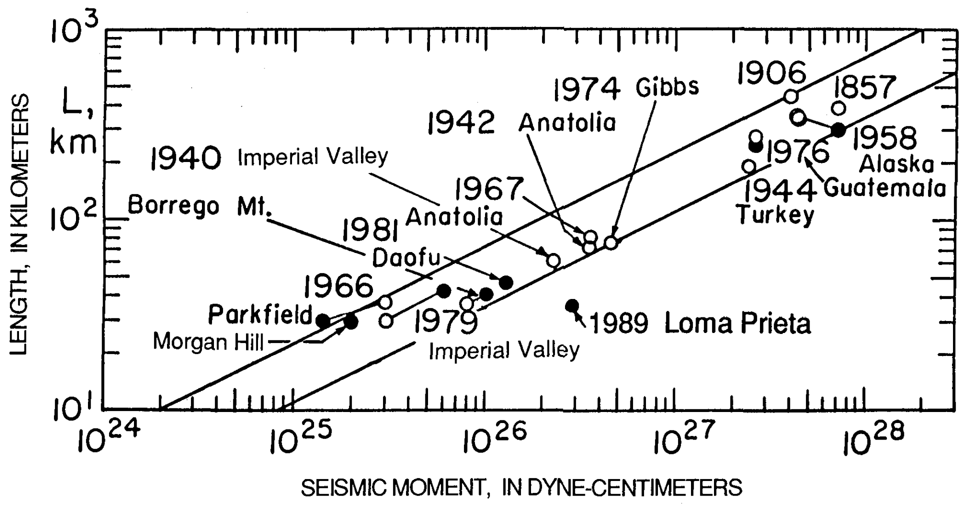

Figure 4.—Fault length as a function of seismic moment for shallow crustal earthquakes along active plate boundaries (from Kanamori and Magistrale, 1989). Solid lines indicate range of data points. Data from Kanamori and Allen (1986) (dots) and Scholz and others (1986) (circles).

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The Loma Prieta, California, Earthquake of October 17, 1989—Main-Shock Characteristics

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